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MOMENTUM FLUX IN OFF-SHORE FLOW

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1. INTRODUCTION

In quasi-stationary atmospheric flow over a homogeneous surface, Monin-Obukhov similarity theory and bulk flux formula with empirical stability functions have successfully predicted turbulent surface fluxes over the sea (e.g., Fairall et al. 1996). Here quasi-stationary means that the total time rate of change of turbulent quantities, such as the turbulence kinetic energy (TKE) or the wind stress, is small compared to production and dissipation terms. However, in regions of significant surface heterogeneity, such as near coastlines, advection by the mean wind and vertical flux divergence can become significant. Then a well-defined surface layer may not exist. In this case, similarity theory will be inadequate due to a dependence of the fluxes on upstream conditions.

With off-shore flow, strong turbulence generated over the relatively rough and often heated land surface is advected over the water where it begins adjusting to the new smoother sea surface and new surface temperature. A conceptual model of this adjustment is internal boundary layer theory. For off-shore flow of warm air over cool water, the residual land-based turbulence advected downwind above the stable IBL is typically stronger than the turbulence associated with the new smoother and colder sea surface. This occurs despite the fact that the residual turbulence generated over land decays through dissipation as it is advected over the water. The stable case is not described by traditional IBL theory (Mahrt et al. 2000).

In this work, we investigate the spatial structure of the mean flow and wind stress in the first 10 *km* off-shore as land-based turbulence is advected over the water, influenced by the new smoother sea surface and modified by the stratification. Our focus is on the stable case of flow of warm air over cool water. Repeated aircraft passes were flown at multiple levels on flight tracks both parallel and perpendicular to the coast. An interpolation technique is applied to develop fetch-height cross-sections that are suitable for future work on verification of numerical models. These case studies provide challenging situations for future attempts to model off-shore flow.

2. DATA

This study analyzes observations from the Atlantic coast near Duck, NC, USA on the Outer Banks (Figure 1) during 2 - 18 March 1999 and 11 November - 4 December 1999 during the Shoaling Waves Experiment (SHOWEX) (Sun et al. 2000). The NOAA LongEZ (N3R) aircraft measured the three components of the wind, air temperature, humidity, surface temperature and atmospheric pressure. The aircraft flight patterns considered in this work include repeated tracks close to and parallel to the coast for a sequence of off-shore distances and heights, as well as repeated tracks perpendicular to the coast for a sequence of different altitudes. The same flight track was flown numerous times to reduce random flux sampling errors.

Turbulence was defined to include all fluctuations on length scales from 1 *km* down to the resolution of the measurements (approximately 1 *m* based on a 50 Hz sampling rate and an aircraft ground speed of 50 ms^{-1}). The flux averaging length scale was chosen as 1 *km* based on a compromise between a length scale large enough to capture most of the turbulent flux yet small enough to resolve the spatial variability.

For flights with adequate spatial coverage and off-shore flow, we interpolate the 1 *km* mean estimates from the aircraft onto a fetch-height (x, z) grid of resolution $\Delta x = 500$ m and $\Delta z = 25$ m with a domain equal to 0 to 10 *km* fetch and 10 to 310 *m* in height (Vickers et al. 2000). Fetch is the distance over the water that an air parcel has traveled since leaving land. The interpolation method weights observations inversely proportional to the distance between the observation and the grid point and ignores small time differences.

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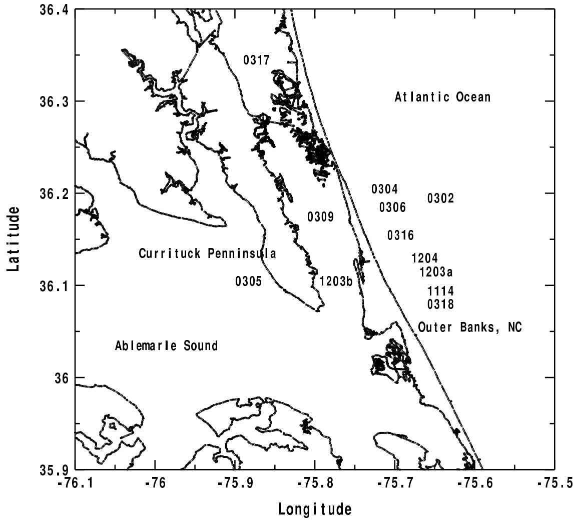

Figure 1: Shoaling Waves Experiment (SHOWEX) study area near Duck, NC, USA, and the location and date (mmdd) of the twelve off-shore flow case studies.

The kinematic sea surface momentum flux (u_*^2), which represents the air-sea exchange of momentum, was estimated by linearly extrapolating the interpolated estimates at heights of 10 and 35 *m* to the surface. In stable conditions, where the momentum flux typically increases with height close to the coast, the extrapolation yields an estimate of u_*^2 that is smaller than the estimate at the lowest aircraft measurement height.

The spatial gradients in the equation of motion for the mean wind,

$$\frac{\partial U}{\partial t} + U\frac{\partial U}{\partial x} + W\frac{\partial U}{\partial z} + \frac{\partial \overline{w'u'}}{\partial z} + \frac{\partial \overline{u'u'}}{\partial x} = -\frac{1}{\rho}\frac{\partial P}{\partial x}, \quad (1)$$

were evaluated from the interpolated fetch-height cross-sections using finite differences evaluated across a conceptual grid box extending from 1 to 3 *km* in fetch and from 10 to 135 *m* height above the water. Horizontal gradients were computed as the difference of the vertical averages over the grid box, and vertical gradients were computed as the difference of horizontal averages. The local time rate of change term was estimated from the earliest and latest aircraft passes for the day, which were typically 1 hour apart. Terms involving horizontal gradients in the direction parallel to the coast were neglected.

The mean vertical velocity (W) in the vertical advection term was estimated by first removing the flight-averaged vertical velocity for each flight day. After removing the flight-averaged vertical velocity, a surprisingly consistent pattern was observed in which the mean vertical motion was downward (upward) in the region of acceleration (deceleration) of the horizontal mean wind.

The horizontal pressure gradient is not known but was crudely estimated as a residual assuming balance with the remaining terms (Eq. 1). Attempts to calculate the horizontal pressure gradient by removing the altitude-dependence of the pressure as measured by the aircraft were deemed unreliable. An estimate of the horizontal pressure gradient due only to the horizontal temperature gradient was made using the hydrostatic equation and the observed temperature field. An estimate of the large-scale horizontal pressure gradient was calculated from the NCEP/NCAR Reanalysis surface pressure (Kalnay et al. 1996).

In traditional turbulence scaling arguments, the turnover time for a large eddy is proportional to L_s/σ_w, where σ_w is the standard deviation of the vertical wind defined for some suitable averaging time, and L_s is the length scale where 50% of the total variance of the vertical wind occurs at smaller scales. To estimate this length scale for the land-based boundary layers, we applied multi-resolution decomposition to the vertical velocity data obtained in the boundary layers over land. The spectra from repeated flight tracks in the same region on the same day were composited by averaging the estimates of L_s for individual passes. The location of the peak of multi-resolution spectra in the frequency domain depends primarily on the scale of the fluctuations, while the peak of Fourier spectra depends on the periodicity.

3. RESULTS AND CONCLUSIONS

A consistent pattern was observed for the stable off-shore flow cases (e.g. Figures 2 and 3). For stable flow of warm air over cool water (air-sea temperature difference between 4 and 8 *C*), the low-level momentum flux over the sea decreases rapidly with increasing fetch for the first few kilometers and gradually reaches equilibrium values by 10 km off the coast. This decrease is thought to be associated with decay of advected land-based turbulence by the mean flow, which accelerates downstream from the coast. Beyond a kilometer off-shore, an increase in the downward momentum flux with height is observed in the lowest 150 m. The initial formation of an elevated wind stress and TKE (not shown) maxima is attributed in part to the height-dependence of advection. For smaller stable air-sea temperature differences (< 2 *C*) and for unstable flow of cool air over warm water, this structure was not observed. In the near-neutral and unstable cases, the mean wind does not significantly accelerate over the sea and the wind stress over the sea decreases with height.

Horizontal advection is the largest term close to the coast in the equation of motion for the mean wind (Eq. 1) due in part to the generally strong wind speeds. The

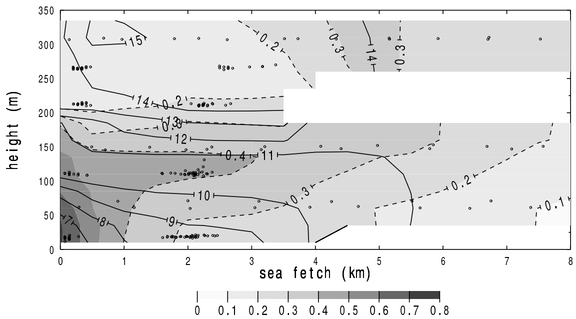

Figure 2: Fetch-height cross-section for case 0318 of the friction velocity (shaded and dotted contours, ms^{-1}) and the mean wind speed (solid contours, ms^{-1}. Dots indicate locations of the 1 *km* data points. Clusters of points result from parallel flight tracks.

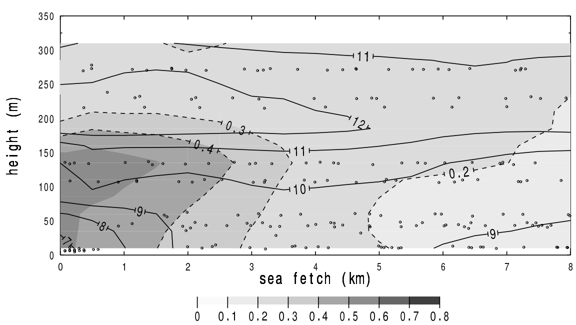

Figure 3: Same as Figure 2 except for case 1114.

vertical advection term and the vertical flux divergence term both act to balance horizontal advection. That is, acceleration of the mean wind in the first few kilometers off-shore is associated with an increase in the downward momentum flux with height and with mean sinking motion. Vertical advection brings stronger momentum down towards the surface since the mean wind speed increases with height. The three cases with these characteristics (0318, 1114 and 0317) are stable flow of warm air over cooler water. The single case of flow of cool air over warm water (1203a) is associated with deceleration of the mean wind, a decrease in the momentum flux with height and mean rising motion. The normally neglected variance of the mean horizontal wind ($\overline{u'u'}$ in Eq. 1) decreases with increasing off-shore fetch, contributing to acceleration of the mean wind. This term is an order of magnitude smaller than the largest term, but not entirely negligible.

The hydrostatic part of the horizontal pressure gradient due to the horizontal temperature gradient is small compared to the residual pressure gradient except in the unstable case. In the unstable case (1203a), the hydrostatic part of the horizontal pressure gradient and the residual horizontal pressure gradient are approximately equal and Eq. (1) is balanced. The large-scale pressure gradient from the Reanalysis model is small in the unstable case. For the stable cases, the large-scale horizontal pressure gradient from the Reanalysis has approximately the same magnitude as both the vertical advection and the vertical flux divergence terms, and is of the correct sign to help balance the horizontal advection, but significant residual remains in the budget. The momentum budget inbalance as a percentage of the horizontal advection term is 30%, 53% and 2% for the stable cases 0318, 114 and 0317, respectively. The inbalance in the stable case may be partly due to errors associated with inadequate sampling and difficulties estimating gradients.

3.1 Advective-decay time scale

Assuming that within the first several kilometers downwind from the coast, the local generation of wind stress over the sea is small compared to the influence due to advection of strong turbulence from land, we can estimate an advective-decay rate. The advective-decay rate refers to the rate of decrease in the sea surface momentum flux with travel time from the coast. From a Lagrangian viewpoint, the advective-decay rate in stable off-shore flow is at least partly due to viscous dissipation of the Reynolds stress. We exclude from consideration cases with convective conditions over the water where local generation of turbulence due to buoyancy effects is not negligible compared to advection. Analogous to traditional scaling arguments for the decay of turbulence variances, and based on the actual travel-time dependence of the sea surface wind stress (Figure 4), we formulate the surface stress as a function of travel time t from the coast as

$$u_*^2(t) = u_{*o}^2 exp(-t/\tau) + u_{*eq}^2(1 - exp(-t/\tau)) \qquad (2)$$

where τ is the advective-decay time scale, u_{*eq} is the equilibrium value of u_* for long travel time over the water beyond the influence of advection from land, and u_{*o} is the initial value at the coast (u_* at $t = 0$). The advective-decay time scale was estimated from data by applying linear least-squares regression to the $u_*^2(t)$ values, of which there is one estimate every 500 m of fetch.

A second, independent estimate for the time scale that is based only on the vertical velocity fluctuations in the upstream boundary layer is

$$\tau_L = (L_s + c)/\sigma_w. \qquad (3)$$

L_s is the eddy length scale for the turbulence in the land-based boundary layer based on the vertical velocity spec-

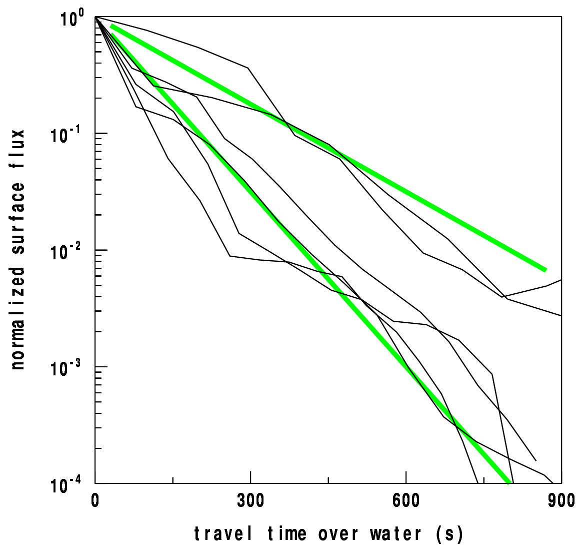

Figure 4: Decrease in the normalized sea surface momentum flux $((u_*^2(t) - u_{*eq}^2)/(u_{*o}^2 - u_{*eq}^2))$ versus travel time over the water (t) for six case studies. The two heavy lines are the predicted decrease from Eq. (2) for values of the advective-decay time scale τ equal to 200 s (lower line) and 400 s (upper line).

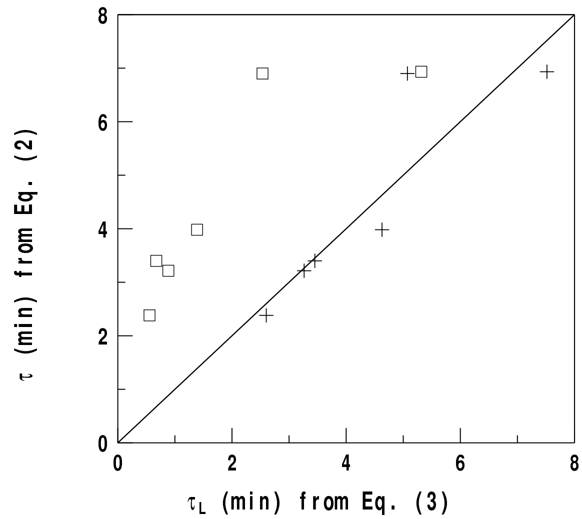

Figure 5: Advective-decay time scale τ based on observed decrease in u_*^2 with travel time (Eq. 2) versus τ_L calculated from characteristics of the upstream land-based boundary layer (Eq. 3 with $c = 0$) for six case studies (squares). The plus signs show τ versus τ_L using $c = 70$ m in calculating in τ_L.

tra, σ_w is the standard deviation of the vertical velocity in the land-based boundary layer and c is an unknown constant. This formulation for the time-dependence of the momentum flux is analogous to traditional theory for dissipation of turbulence variances, where the dissipation time scale increases with the characteristic length scale of the eddies and decreases with the strength of the mixing.

The time scale τ, calculated from the observed rate of decrease in u_*^2 (Eq. 2), and the time scale τ_L, based on upstream conditions only (Eq. 3 with $c = 0$), are related (Figure 5). This relationship implies that the sea surface momentum flux near the coast in stable off-shore flow depends on the characteristics of the boundary layer over land. Using an alternate eddy length scale for the turbulence in the land-based boundary layer equal to L_s plus a constant ($c = 70$ m in Eq. 3) leads to a linear one-to-one relationship between τ and τ_L (Figure 5). A possible interpretation of $c > 0$ is that the turbulence is not in pure decay and that generation of turbulence over the water is not negligible.

4. ACKNOWLEDGMENTS

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